

Collaborative Research: Climate Sensitivity of Thaw Lake Systems on the Alaska North Slope

(Final Report)

Principal Investigator: Tingjun Zhang, University of Colorado at Boulder,
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This was a collaborative research project with leading institution of the Geophysical Institute, University of Alaska Fairbanks and Principal Investigator, Dr. Martin O. Jeffries.

The main **objective** of the proposed study was to understand the present day ice/lake/soil interactions and thermal regime of shallow tundra lakes on the North Slope of Alaska, and to investigate their sensitivity to climate change, in order to determine the role of the thaw lakes in permafrost development and distribution and potential changes in greenhouse fluxes from the tundra. The proposed research will combine remote sensing of lake ice processes with two-dimensional numerical modeling of ice/lake/soil interactions and thermal regime. A number of individual objectives was addressed:

- 1). quantify the spatial variability of the number of lakes that freeze completely to the bottom each winter and those do not;
- 2). quantify the present thermal regime of the thaw lakes and the underlying and adjacent soils, including the extent of lateral and downward freezing and thawing at lakes that freeze completely to the bottom and those do not; and
- 3). model the effects of climate change (warming/cooling) on ice growth and the number of lakes freezing completely to the bottom, and the consequences for the thermal regime of the lakes/soils.

The main task for the University of Colorado team was to develop the two-dimensional heat transfer model with phase change to simulate the ice/lake/soil interactions, thermal regime of permafrost under lakes, and potential response of lake ice growth and permafrost thermal regime under thaw lakes to climatic change. This report is mainly focused on the University of Colorado main task and accomplishments.

Introduction

There are thousands of thaw (thermokarst) lakes on the North Slope of Alaska, where they cover as much as 40% of the land area. Their very name recognizes the fact that they owe their origin to the impact they have on the ground thermal regime, but there have been few quantitative studies of the impact of the lakes on atmosphere-land interactions in this tundra region.

Model Development

We developed a physically-based, non-steady heat transfer model with phase change using a finite element method to investigate heat and mass transfer between the

atmosphere and permafrost, through the intervening snow cover, lake ice and lake water. Detail description of the model is given by Zhang and Jeffries (2001) and a brief introduction will be provided below.

This finite element numerical model is based upon the originals of Guymon and Hromadka (1977) and Guymon and others (1984), which was later modified by Gosink and Osterkamp (1990). The model is assumed that there is no convective heat flow in the system. It is generally true that heat transfer in lake ice is dominated by conduction. Thaw lakes in northern Alaska are generally shallow, mostly ranging from 1 to 3 m (Sellmann and others, 1975a, 1975b; Mellor, 1982; Jeffries and others, 1996). Water under the lake ice in winter or lake water in summer are usually well mixed and close to isothermal. Therefore, convective heat transfer is relatively small or negligible.

This model is coupled with two moving boundaries: (i) air–snow interface as snow thickness changes; and (ii) the phase boundary between lake ice and lake water as lake ice grows and melts. The seasonal snow cover is treated as an additional layer when it is present. The effective thermal properties of snow are estimated using a serial heat conduction model (Zhang and others, 1996), which includes wind slab and depth hoar layers as reported by Benson and Sturm (1993). The density of snow over lake ice is obtained from Jeffries and others (1999).

The model includes up to six different types of materials, individually or in combination, such as snow, ice, water, peat, silt, sand, and gravel. The material type can be varied from layer to layer with distinct physical and thermal properties. The values of the thermal properties of the materials (snow, ice, or water) were determined during the calculations, depending on the position of the phase boundaries. Latent heat is included as a latent heat content “spike” at the given freezing temperature (O’Neill, 1983). Lake ice melting is a more complicated process than lake ice growth. During the decay period of lake ice, both the top and the bottom of the floating lake ice cover are at 0°C. Lake ice usually is isothermal at 0°C within a few days after the onset of melting at the top, but the complete melting of lake ice takes a few weeks or more than a month in the Arctic and sub-Arctic (Bilello, 1980). In this case, heat conduction is not applicable. Bilello (1980) presented extensive data from a variety of locations in the Arctic and sub-Arctic and examined several empirical approaches to predict the decay rate of lake ice. The accumulated thawing degree-day (ATDD) index gave the best overall correlations, especially when the lake ice cover was not subjected to water currents or other action that would mechanically break up the ice cover, which is the case near Barrow, Alaska in this study.

In the current model, all ice growth occurs by congelation at the base of the ice cover. However, lake ice also thickens by the formation of snow ice, particularly in sub-Arctic locations. Congelation ice and snow ice are also commonly referred to as black ice and white ice, respectively, in the literature (e.g., Adams, 1981). Snow ice forms after the snow/ice interface has been flooded, typically by water upwelling through cracks when the snow load is sufficient to depress the snow/ice interface below lake level. We have observed snow ice in the ice cover on lakes in the Barrow area, but its occurrence tends to be patchy and the layers thin, and most of the ice cover is composed of congelation ice (Jeffries and others, 1994). Since snow ice makes up only a small proportion of the total

area and mass of lake ice, we consider the simulations to be realistic as they represent the dominant (area and thickness) ice type, i.e., congelation ice.

Major Results

- The melting factor of lake ice near Barrow, Alaska, is about $1.53 \text{ cm/}^{\circ}\text{C-day}$, slightly higher than the value of $1.29 \text{ cm/}^{\circ}\text{C-day}$ at Barter Island, Alaska and significantly higher than values obtained in the 1970s in northern Canada.
- The long-term (1947–97) mean freezing index (from July 1 through June 30) is $4824 \text{ }^{\circ}\text{C-day}$, with extremes varying from $5708 \text{ }^{\circ}\text{C-day}$ to $4180 \text{ }^{\circ}\text{C-day}$. The long-term mean snow cover index is cm-day with extremes ranging from 10447 cm-day to 2040 cm-day . The simulated long-term mean maximum lake ice thickness is 1.91 m with extremes of 1.33 m to 2.47 m .
- Change in the seasonal snow cover is the most important single factor controlling inter-annual and inter-decadal variations of lake ice thickness in northern Alaska. Winter monthly air temperature or freezing index is an important but secondary factor influencing variations of lake ice growth and thickness.
- There is a strong anti-correlation between the snow cover index and maximum lake ice thickness, i.e., maximum lake ice thickness is greater during years with lower snow cover index than during years with higher snow cover index.
- Maximum lake ice thickness was a little above its long-term average during the mid-1950s but decreased substantially during the mid-1960s because of the combined impact of higher snow cover index, and early and greater snowfall, even though the freezing index was well above its long-term average. Lake ice thickness reached peak values during the mid-1970s mainly due to the lower snow cover index, colder winters (higher freezing index), and late snowfall. Subsequently, lake ice thickness generally decreased due to the increased snow cover index and decreased freezing index.
- Sensitivity studies indicate that a change in maximum snow depth from 0.0 cm (no snow) to 0.75 m would decrease maximum lake ice thickness from 2.62 m to 0.69 m , a 73% decrease. A change in the bulk thermal conductivity of snow from $0.08 \text{ Wm}^{-1}\text{K}^{-1}$ (fresh snow) to $0.40 \text{ Wm}^{-1}\text{K}^{-1}$ (wind-packed snow) would increase maximum lake ice thickness from 0.97 m to 2.07 m , a greater than two-fold increase. A change in the freezing index of air temperature from 5781°C-day to 3909°C-day would decrease maximum lake ice thickness from 2.10 m to 1.62 m , a 23% decrease.
- For extreme cases, maximum lake ice thickness decreases from 2.68 m with model inputs of minimum daily air temperature and snow depth, to 0.99 m with model inputs of maximum daily air temperature and snow depth, a 63% decrease.

This study indicates that maximum lake ice thickness varies significantly from year to year in northern Alaska. A recent study indicates that the mean conductive heat flux through snow cover on floating lake ice (ice is not frozen to the bottom) is about 2.4

times greater than the mean heat flux through snow cover on grounded ice (ice is frozen to the bottom), and 11.4 times greater than the heat flux through snow cover on tundra (Jeffries and others, 1999). Changes in maximum lake ice thickness could have a direct impact on the area of ice and the numbers of lakes which do and do not freeze to the bottom each winter in the region, considering that most of the thaw lakes in northern Alaska are 1 to 3 m deep (Mellor, 1982). Consequently, changes in maximum lake ice thickness would have a significant impact on the regional energy balance, and thus a significant feedback to the climatic system.

The minimum condition for the formation of a talik (thaw bulb) is that lakes do not freeze to the bottom by the end of winter. Changes in maximum lake ice thickness would have a significant effect on the thermal regime of permafrost and inter-annual and inter-decadal talik formation at lakes where water depths vary within the range of maximum lake ice thickness over the past 50 years (Figure 2c). For example, for lakes with a water depth of 1.91 m, the long-term average maximum lake ice thickness, a talik may form during years when the lake does not freeze to the bottom and may refreeze during years when the lake freezes completely to the bottom. Talik formation and refreezing under thaw lakes in northern Alaska would have a significant impact on methane release from lakes to the atmosphere. For example, during years when a talik forms under thaw lake, the methane efflux may be substantially greater when permafrost thaws than during years when the talik refreezes. From the simulated time series of the inter-decadal variations of maximum lake ice thickness (Figure 2c), we may speculate that the methane efflux might be greater over thaw lakes during the late 1950s and the 1960s, when maximum lake ice thickness was relatively low, than during the 1970s, when maximum lake ice thickness reached peak values.

This study demonstrates that lake ice growth and thickness can be modeled with confidence, provided the necessary field measurements available. However, the currently available data are spatially and temporally sporadic. This study and others indicate that lake ice thickness can vary substantially from year to year, and from place to place, due to the spatial and temporal variations of environmental conditions, especially snow cover parameters. Systematic field measurements, model improvements and validation are essential for better understanding the thaw lake system and for further investigation of the inter-decadal variations of lake ice growth and thickness in the Arctic and sub-Arctic.

Simulations show, for example, that a 3 m deep lake with 400 m radius will develop a 50 m deep talik after 3200 years. Simulations also show that only those lakes > 2m deep develop a perennial talik (thaw bulb), regardless of the age of the lake. This indicates that the long-term mean maximum seasonal ice thickness is 2 m; hence, any lake < 2 m deep will freeze completely to the bottom each winter and only a seasonal thaw zone will develop each summer and refreeze each winter.

SAR data have been used to validate the onset of snow melt and the decay of the ice cover, as simulated by the model. SAR data were also used to determine the area of tundra, floating ice and grounded ice (frozen to the lake bed) in mid-April 1997 when the conductive heat flow from the lake ice and tundra to the atmosphere was derived from field measurements. The SAR and field data were combined to estimate the area-

averaged heat flow ($4\text{--}5 \text{ W m}^{-2}$) to the atmosphere from the North Slope, and it was found to be equal to that from the nearby Arctic Ocean. Simulations of the conductive heat flow from the lakes and tundra to the atmosphere during winter 1996–1997 revealed a maximum of 35 W m^{-2} in early winter. It is concluded that the heat flow from these lakes might have a significant effect on the local climate.

The SAR mapping of the area of floating and grounded ice in mid–April 1997 led to the development of a method to map the bathymetry of the shallow ($<2 \text{ m}$) zones of thaw lakes. The mapping method relies on the fact that the boundary between floating ice and grounded ice is easily identified in SAR images, and that the date a particular SAR image is acquired can be converted into ice thickness using a simulated ice growth curve. The simulated ice thickness represents the water depth at the boundary between floating and grounded ice. A sequence of SAR images is used to produce a map of irregularly spaced isobaths, and subsequent interpolation and smoothing is used to produce a final product with regularly spaced isobaths, e.g. at 0.25 m intervals.

Peer-reviewed Publications

Zhang, T. and M. O. Jeffries. 2000. Modelling inter-decadal variations of lake ice thickness and sensitivity to climatic change in northernmost Alaska. *Annals of Glaciology*, 31, 339–347.

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Conference Presentations

Zhang, T. and M.O. Jeffries, 1999. Thermal and hydrological regime of lakes in a tundra environment along the Alaskan Arctic coastal regions. AGU Fall Meeting, San Francisco, December 1999.

Zhang, T. and M. O. Jeffries, 1999. Modelling inter-decadal variations of lake ice thickness and sensitivity to climatic change in northernmost Alaska. International Glaciological Society Symposium on The Verification of Cryospheric Models: Bringing Data and Modelling Scientists Together, Zurich, Switzerland, 16–20 August 1999.

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Jeffries, M. O., T. Zhang, K. Frey and N. Kozlenko, 1998. Contribution of Shallow, Thaw Lakes to the Winter Heat Flux from the Alaskan North Slope. AGU Fall Meeting, San Francisco, December 1998.

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Zhang, T., K. Stamnes, and W. Zhou, 1997. Influence of the seasonal snow cover on the active layer and permafrost in the Arctic and Subarctic, Association of American Geographers 93rd Annual Meeting, Fort Worth, TX, 1-5 April 1997.

Zhang, T., K. Stamnes, and W. Zhou, 1996. Approximation of soil heat fluxes with freezing and thawing suitable for GCMs. AGU Fall Meeting, San Francisco, December 1996.